

# Soil Characteristics Influencing Nutrient Supply in Forest Soils

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## I. INTRODUCTION

Forest plantations occur on almost all types of soils, ranging from nutrient-poor mining wastelands to nutrient-rich vertisols. On a world scale, all the ten soil orders of the United States soil classification system are associated with major forest ecosystems (Pritchett, 1979). Although the number of soil orders represented in a region may be limited, there may be many soil units possessing different soil characteristics, and thus demanding varied management practices. For example, Raupach (1967) named eight important soil types in a small area under *Pinus radiata* (radiata pine) plantations in South Australia: ground water podzols, podzols, yellow podzolics, red podzolics, lateritic podzolic soils, meadow podzolics, krasnozems and terra rosas. The nutrient-supplying capacity, and consequently the productive potential of these soils varies, being high for the krasnozems and the terra rosas and low for the podzols.

In the process of land planning on a regional scale for plantation forestry, variations in soil types could be used as one of the criteria for the selection of proper tree species as shown by Otto (1972) for Lower Saxony, West Germany. The nature of the soil parent material is usually the most important determinant of soil physical, chemical and biological characteristics, and this is often the main parameter used in the selection of sites for plantation forestry (Nakos, 1979). The influence of parent material is reflected through buffering systems in the soil (see Section IV).

Traditionally, plantation forestry involved the use of slow-growing species (i.e. long rotations) and a low intensity of wood utilization (mainly use of stem wood). Because of increasing demand for wood resources, management is now more intensive, involving the use of fast-growing species, shortened rotations and more intensive utilization (e.g. whole tree harvesting, utilization of thinnings). These changes in management practice are likely to increase the drain on the nutrients from already-impoverished sites; e.g. Crane and Raison (1980) calculated a 3–5-fold increase in the removal of P when sites growing eucalypts are converted to pines. Enhanced productivity results in greater demands on the nutrient-supplying capacities of soils. Not only are greater amounts of nutrients needed, but they must be absorbed more rapidly, so soils must contain ample nutrients in readily available forms. Successful intensive utilization of forested lands thus depends to a great extent on the proper management of soil nutrients. This involves a proper understanding of relationships between the chemical characteristics of plantation soils and the soil processes relating to nutrient supply.

The chemical characteristics of plantation soils affect tree growth by influencing not only the soil processes relating to nutrient supply but also

other ecosystem processes such as litter decomposition, microbial and faunal activities. Nutrients move in the ecosystem mainly in ionic form. Their mineralization during decomposition of litter leads to the appearance of ions in the soil solution. Their uptake by plant roots or micro-organisms occurs also as ions. During the movement of ions from one phase to another, e.g. from organic matter to soil solution (mineralization), electrical neutrality must be maintained in each phase. Ions such as  $\text{Na}^+$  and  $\text{Cl}^-$ , which are usually not considered as important nutrients, may therefore play a significant role in nutrient uptake by maintaining neutrality. Imbalance of cations and anions in the process of either mineralization or uptake leads to the production or consumption of protons. The soil is the reaction vessel in which both processes are coupled. Discoupling of mineralization and uptake may, in the long run, lead to the acidification (podzolization) of soils. This is a very important factor in the long-term stability of forest ecosystems, and is discussed in Section V.

Any change in soil chemical characteristics, especially one leading to acidification, would influence the supply of nutrients to roots (Ulrich, 1981a). In acid soils toxic ions, especially aluminium ions ( $\text{Al}^{3+}$ ), appear in the soil solution and influence plant growth. Aluminium plays an important role in terrestrial ecosystems because of its abundance in soil minerals (aluminosilicates), because of its tendency to go into solution at low pH, and because of its toxicity if it enters the living cells of micro-organisms and plants in ionic form. Some of these relationships between soil chemical characteristics and proton buffer systems are discussed in Section IV.

This chapter deals with the chemical characteristics of forest soils, with emphasis on those under plantations, with the intention of providing a basis for understanding the nutrient-supplying capacity of soils in relation to nutrient demands by plantation trees. Changes in soil chemical characteristics as a consequence of management practices, including the introduction of exotic tree species in plantations, are considered in relation to buffering systems operating in different soils. A theory on the stability of forest ecosystems based on nutrient dynamics is presented. Based on these theoretical considerations, silvicultural practices for a sound management of plantations are discussed.

## II. NUTRIENT POOLS AND DYNAMICS

The form in which plant nutrients occur in the soil affects both the rate at which they come into equilibrium with the soil solution (a pre-condition for

their uptake by trees) and the amount in the soil solution. A generalized compartmentalization of nutrient pools in soils may be shown thus:

soil solution  $\rightleftharpoons$  easily-mobilizable  $\rightleftharpoons$  mobilizable with difficulty  $\rightleftharpoons$  immobilizable

Mobilization means that ions are released from solid phases and enter the nutrient pools associated with plants. Exchange among the different pools occurs mainly via the solution phase, which at low soil moisture contents may be reduced to water films. The mobilization reaction may be a reversible chemical equilibrium or an irreversible enzyme-catalysed reaction associated with the mineralization of organic matter. Easily mobilizable pools rapidly reach relatively high solution concentrations; pools mobilizable with difficulty either reach relatively low solution concentrations or react slowly.

A seasonal cycle in the nutrient dynamics of forest ecosystems may exist, controlled by temperature in cold regions and by water availability in the more arid regions of the world (Van Cleve, 1967). Even-aged plantations show nutrient changes synchronous with the forest rotation, and there may also be unidirectional development lasting several rotations.

For short-term seasonal uptake of nutrients there must be relatively fast nutrient supply and hence rapid mobilization of nutrients. This involves, primarily, the soil solution pool and that part of the easily mobilizable pool which stays in rapid equilibrium with the solution pool, for example salts with a solubility comparable to or higher than  $\text{CaCO}_3$  (bicarbonates, nitrates, chlorides, sulphates), the exchangeable cations (Na, K,  $\text{NH}_4$ , Mg, Ca; in acid soils also Mn, Al, Fe), and anions (phosphates) bound on surfaces. The involvement of organically-bound nutrients will depend solely upon the rate of mineralization, which in turn depends upon the characteristics of the substrates (e.g. C/N, C/P ratio) and the climatic conditions which determine the activities of the decomposer organisms. The short-term supply of nutrients to plants thus depends upon physico-chemical processes as well as microbial activity in soils.

The requirements for nutrients by trees over the whole rotation is affected to a significant extent by the weather and other episodic events (insect attack, disease, windthrow and silvicultural practices: thinning, pruning, fire etc.) which influence not only the growth patterns of trees but also nutrient mobilization in soils. In this case the type of nutrient pools involved are hydrolysable N fractions, inorganic P fractions, organic forms of P and S, exchangeable cations and cations held on specific exchange sites.

The long-term nutrient requirements of plantation forestry extend over successive rotations, during which management practices such as changes in land use, forest type, site preparation and harvesting can severely affect the nutrient supply. For a *P. radiata* stand, Squire and Flinn (1981) calculated

that during a 40-year rotation with four thinnings, whole-tree logging, a hot broadcast burn or windrow heap and burn, complete cultivation, planting and vigorous chemical weed control, the combined loss of N would approach  $1380 \text{ kg ha}^{-1}$  or 39% of the total N in the whole system (soil depth 50 cm) on a podsolized sand in South Australia. Some part of the mobilizable pools may be lost due to soil erosion, raking and removal of litter and leaching to deeper soil depths. On these time scales accretion of some elements from the atmosphere becomes significant (see Chapter 3) (particularly in coastal areas, and areas affected by man-made emissions), as does the accumulation of nutrients by weathering of primary silicates. (Of special significance are soils and soil horizons showing silicate buffering range; see Section III, D.)

#### A. Assessment of Nutrient Pools and Turnover

Assessment of the type and size of nutrient pools requires application of chemical analytical methods to soil samples. Such an approach is subject to sampling problems, to chemical problems in defining different types of nutrient pools, and to analytical problems (see Chapter 10). The last of these will not be considered here, and the chemical problems only to the extent that such consideration is useful in understanding the ecological principles operating in plantations.

Sampling problems include those of determining the number of samples required to describe a site, the relevant depth of sampling, the appropriate time of sampling, and the method of sample preparation for analysis. Beckett and Webster (1971) observed that variances and coefficients of variation (CV) of soil properties increased somewhat with the size of the area sampled, but that up to half of the variance within an agricultural field may be present within  $1 \text{ m}^2$ . This suggests that variability may be mainly caused by the processes connected with ion uptake and decomposition. Beckett and Webster found median CV values to be around 35% for organic matter content and 58% for exchangeable cations within the topsoil of a soil series. However, Wright and Wilson (1979) observed that when dealing with a large area the variance does not necessarily increase with the size of the sample area.

Lateral variability in soil properties is particularly high under forests (Kang and Moorman, 1977) due to the heterogeneous nature of the forest floor (Stutzbach *et al.*, 1972) and soil horizons (Hart *et al.*, 1969) in terms of organic matter content, stone and rock content, and distribution of roots. Tree size is always variable, even in plantations, which may lead to variations in the accumulation of litter (Hamilton, 1972), in the rate of water

intake and percolation (Warrick and Nielsen, 1980) and in the microsite climate and other biotic factors which affect mineralization (Lodhi, 1977; Keeney, 1980). Ulrich (1971) observed, for a 1 ha area under a 120-year-old *Fagus sylvatica* (beech) stand on acid brown earth in northwest Germany, CVs of 2% in soil pH and Schofield's potentials (negative logarithms of activities of cation ratios in the equilibrium soil solution), 5–10% for C/N ratio and phosphate pools, 15–20% for C- and N-pools, and 6–60% for pools of exchangeable cations—the highest value being for exchangeable Na. Spatial heterogeneity can sometimes make it difficult to fulfil statistical requirements within practical restraints on sampling. One way to overcome the problem of keeping the number of samples low is to bulk the individual samples, or to collect randomized stratified samples. In many instances the latter will be the only choice if treatment effects (e.g. burning of forest residues) are highly variable and/or are small compared to the variability on the site.

The depth to which soils should be sampled is another important consideration. In principle sampling should include the feeding zone of roots (Voigt *et al.*, 1964). A homogeneous distribution of roots with depth might reflect relatively homogeneous soil physical and chemical characteristics. However, the distribution of fine roots and thus of soil chemical characteristics usually follows steep gradients with increasing soil depth, and such gradients may even occur within a single soil horizon. In such cases, calculation of soil nutrient content should depend upon the volume of soil which may be exploited by the tree roots. Voigt *et al.* (1964) calculated the amount of exchangeable Ca and K in the volume of soil within 1 cm of any root surface of ten 5–9 year *Pinus rigida* trees growing on an abandoned gravel pit in New Jersey. The mean values were 198 mg and 35 mg for Ca and K respectively, which compared with 16.4 g and 3.53 g for their amounts in the total rooting zone. This is due partly to the low rooting density of some trees (see Chapter 6).

Time of sampling is important for some soil parameters subjected to short-term seasonal changes in the ionic cycle. For example, based on detailed studies on the pattern and seasonal variation in the amount of nutrients in a *Pinus sylvestris* (Scots pine) forest soil (humus iron podsol), Usher (1970) recommended that soil samples for survey work should be taken in the spring or early autumn, at a standard distance from trees, and should be cores of more than 13 cm diameter. Time of sampling is of interest especially for studies on mineralization of organic matter and for those on assessing pH and other acidity linked parameters in soils showing chemical soil characteristics corresponding to the exchange buffer range (see Section IV, C).

The fine soil fraction (<2 mm) is usually used for chemical analysis. Therefore the assessment of nutrient pools on a land-area basis necessitates

collection of information on stone content, bulk density and root distributions with soil depth as well as changes (induced by management practices) in these factors. The amount of nutrients per hectare can be calculated from:

$$\begin{array}{l} \text{Amounts of} \\ \text{nutrients} \\ (\text{kg ha}^{-1} \text{ depth}^{-1}) \end{array} = \sum_{\text{soil layers}} \left( \begin{array}{l} \text{Layer} \\ \text{thickness} \times \\ (\text{dm}) \end{array} \right) \times \begin{array}{l} \text{Bulk density} \\ (\text{g cm}^{-3}) \end{array} \times \begin{array}{l} \text{Fraction of} \\ \text{fine earth} \end{array} \times \begin{array}{l} \text{Element} \\ \text{concentration} \\ \text{in fine earth} \\ \text{fraction} \\ (\text{mg kg}^{-1}) \end{array}$$

where fraction of fine earth =  $1 - \text{stone } (> 2 \text{ mm}) \text{ fraction}$

A good estimate of bulk density is very important for assessing the size of nutrient pools in soils. The values may vary from  $0.2 \text{ g cm}^{-3}$  for litter or humus layers to  $1.8 \text{ g cm}^{-3}$  in compacted sandy horizons in a subsoil. For surface horizons (0–5 cm) under woodlands in the English Lake District, Harrison (1979) observed the following bulk densities for different humus types: 0.26 for mor, 0.36 for moder and 0.61 for mull. The bulk density of surface soils may vary according to the stage of the tree rotation (Lundgren, 1978) depending upon the physical characteristics of soils and the method of conversion of native bush to plantations. For example, Hamilton (1965) observed an increase of 7–20% in bulk density of surface soils when planted with *Pinus radiata* following eucalypts; this was associated with a decrease of 37–60% in organic matter content. Soil sampling for diagnostic purposes is discussed in Chapter 10.

Soil testing methods seldom give an insight into the chemical form of elements in the soil, and thus do not provide the basis for understanding the processes affecting nutrient uptake by trees (see Chapter 10). A useful approach, which will be used here, is that of dividing the total amount of soil nutrients into fractions having different rates of mobilization/immobilization, solubilization/precipitation, fixation etc. This allows an understanding of how elements move from one form to another in response to changing conditions of soil and vegetation.

Two approaches might be used to fractionate the total nutrient pools of soils:

- (a) Physical fractionation of the soil into organic matter, sand, silt, clay and the hydrous oxides of Fe, Al and Mn, followed by the determination of the nutrient content in each of the fractions. This type of fractionation has been used for determining the fate of applied fertilizer. To a large extent the resulting fractionation of nutrients depends on the type of soil being sampled. In soils from southeastern United States, Schuman

(1979) observed that those with fine texture had a large proportion of their Zn and Cu in the clay content, but that coarse-textured soils had relatively higher percentages in the organic matter fraction. Because the different physical fractions can contribute simultaneously to nutrients in the soil solution, fractionation of this kind is of limited value in understanding plant nutrition.

- (b) The fractionation of a nutrient into various chemical forms such as those soluble in water, exchangeable, non-exchangeable, organic forms, etc. Such a fractionation may conform to sequence of forms of nutrients which are utilized by plants, and to the sequence in which any added nutrients would be transformed.

### B. Nitrogen Pools in Soils

Most of the natural input of N to forest soils occurs as proteinaceous residues of plant or animal origin, which undergo further transformation into organic or inorganic forms mainly via microbial and faunal activities (see Chapter 5). Organic forms of N usually constitute more than 95% of the total N in organic-rich horizons. In the deeper soil horizons,  $\text{NH}_4^+$ -N may be fixed within clay lattices which may form a significant fraction of total N. About 70–80% of the organic N in the surface horizons of forest soils is hydrolysed by acid (6N HCl), alkali or enzyme (pepsin) to produce  $\alpha$ -amino-N as the main fraction (usually 20–50% of total N). However, when the decomposition of organic matter is restricted in forests with mor types of humus (moder and raw-humus)  $\alpha$ -amino-N fraction may be as high as 50–60% of total N (Wittich, 1952), which may be due to  $\alpha$ -amino-N forming stable complexes with clay minerals in the humus-rich A horizons (Bremner, 1965).

Other fractions of N in the hydrolysate are aminosugars (2–10% of organic N), amide or ammoniacal N (10–20%) and unidentifiable N (15–30%). Because the aminosugars e.g. glucosamine, are mainly of microbial origin, their concentration increases with depth in the humus layer in the order:  $O_I < O_F < O_H < A_h$  ( $I$ ,  $F$  and  $H$  refer to undecomposed, fragmented and humified layers of litter respectively; and  $A_h$  to humus rich A horizon). The same sequence may be observed for amide-N (glutamine, asparagine) which is of microbial or plant origin. The amount of amide-N also changes with the type of the humus: mull > moder > raw-humus. Unidentified hydrolysable N is chiefly composed of amino-N but may also contain derivatives of purines and pyrimidines released from nucleic acids, and other compounds containing N. The nature of the non-hydrolysable N (15–20% of organic N) is not well defined. It is also called heterocyclic bound N, as it is assumed to



contain complexes of phenol-protein. Such complexes are resistant to microbial decomposition (Handley, 1954). Jansson (1958) emphasized that a distinction must be drawn between the biologically active and relatively inert soil N fractions.

$\alpha$ -amino-N and amide-N are readily mineralized in soils. However, in any one season not more than 1–2% of total N in soils is usually mineralized. For example, Rapp *et al.* (1979) observed that under a 35-year-old *Pinus pinea* plantation, out of the 706 kg of total N ha<sup>-1</sup> (69, 197, 196 and 244 kg in F<sub>1</sub>, F<sub>2</sub>, F<sub>3</sub> and A<sub>1</sub> horizons respectively) only 11.5 kg were mineralized in one year (3.7, 4.2, 1.0 and 2.6 kg in F<sub>1</sub>, F<sub>2</sub>, F<sub>3</sub> and A<sub>1</sub> respectively). Due to inherent soil variability it is difficult to detect such small differences in soil-N fractions. However, changes involving cumulative differences over a longer period may be ascribable to individual N forms which are expected to be better indicators than the total N which is usually studied.

Among other approaches to the fractionation of the total N in soils, that recently described by Paul and Juma (1981) requires special consideration. They divided total N into six pools according to their turnover rates; namely, old organic matter, the stabilized fraction, the active fraction, biomass, metabolites and the inorganic forms (NH<sub>4</sub>-N and NO<sub>3</sub>-N). They suggested a way of determining the active N fraction in soils by using <sup>15</sup>N. In their experiment entailing 12 weeks of incubation, 24% of the N mineralized was derived from microbial biomass, 4% from metabolites, 32% from the active N pool and 40% from the stabilized N pool. The transfer rates (day<sup>-1</sup>) of various N pools were: metabolites 1.0, biomass 0.0143, active pool 0.0037 and stabilized and old pool  $6.0 \times 10^{-4}$  to  $3.0 \times 10^{-6}$ .

The mineralization of organic N in forest soils depends upon the chemical composition of the substrate (e.g. C/N, C/P ratios; lignin and cellulose contents), microclimatic conditions in the soil, and the nature of the microbial population (see Chapter 5). The ammoniacal-N, which is formed in this process is the main form of N utilized both by plantation trees (Cole, 1981) and micro-organisms (Jansson, 1958). Only under special conditions will NH<sub>4</sub>-N accumulate in soils or be oxidized to nitrite and subsequently to nitrate (Vitousek *et al.*, 1979). The ratio of NH<sub>4</sub>-N to NO<sub>3</sub>-N in soils is therefore regulated by the nitrification process, the rate of which decreases with increasing soil acidity. In a calcareous mull the ratio was 0.1, in a moder (with low arthropod activity) ~ 1, and in raw humus (with almost no arthropod activity) ~ 10 (Baum, 1975). The amount of NO<sub>3</sub>-N in most forest soils is usually low except when the humus type is mull (Baum, 1975), or when the humus is breaking down after disturbance (e.g. clearcutting, Likens *et al.*, 1977; Vitousek *et al.*, 1979), when the system has received excessive amounts of soluble N fertilizer, or when the system is disturbed by high inputs of N in anthropogenic emissions (Ulrich *et al.*, 1979). Nitrates,

once produced, may be lost from the system either by leaching or denitrification unless they are taken up by roots. The leaching of nitrates from terrestrial ecosystems is affected by different management practices such as harvesting, cultivation, fire, fertilization etc (Khanna, 1981).

Nitrogen status of plantation soils is usually assessed by the rate of mineralization of organic N, obtained usually by *in situ* soil incubation studies (Melillo, 1981; Powers, 1980). The net mineralization rates vary from low values ( $0-24 \text{ kg N ha}^{-1} \text{ yr}^{-1}$ ) in mountainside forests on rendzina soils to high values ( $375-399 \text{ kg N ha}^{-1} \text{ yr}^{-1}$ ) on ravine forests on colluvial braunerde (Melillo, 1981). The mean annual net mineralization rate for zonal deciduous forests of central Europe is approximately  $110 \text{ kg ha}^{-1}$  and compares with the mean annual uptake rate of  $115 \text{ kg ha}^{-1}$  for five European deciduous forests (Melillo, 1981). Yield potential and foliar N content of *Pinus ponderosa* (ponderosa pine) growing on volcanic, metavolcanic and metasedimentary soils could be related to N mineralized during 14-day anaerobic incubation at  $30^\circ\text{C}$  (Powers, 1980). Values obtained in that way compared in magnitude with those obtained on anaerobic incubation for 6 months in the field. Values less than  $12 \text{ g N g}^{-1}$  soil showed that N was deficient. However, such studies to assess the N-supplying status of plantation soils should preferably be done under least disturbed *in situ* mineralization studies following the method proposed by Rapp *et al.* (1979).

### C. Phosphorous Pools in Soils

Total soil P can be partitioned into inorganic forms associated with Al, Fe, and Ca (amounts of Ca-P being insignificant under acid soil conditions), forms occurring as occluded inorganic P, and those found as organically-bound P. They continuously undergo transformations in form by chemical processes (determined by pH buffering ranges; see below and Section IV) and by biological processes (e.g. plant uptake, and microbial decomposition), and translocations in soil profiles by processes such as litterfall, microbial and faunal activity (Walker 1965; Ulrich and Khanna 1969).

#### (1) Inorganic P and its availability to plants

Ulrich and Khanna (1968) observed that soil pH was a major factor determining the relative abundance of different inorganic P forms in 230 forest soils of northwest Germany. However, other factors such as the organic matter content, the soil depth, the amount of  $\text{CaCO}_3$ , waterlogging and fertilization modified the proportions of P forms (Table I). When the

Table 1. Forms of soil phosphorus in some forest soils of northwest Germany grouped according to the lime potential and characteristics of soils.<sup>a</sup>

Lime potential (pH-0.5 pCa)	Other soil characteristics	P fractions					
		n <sup>b</sup>	P <sub>Al</sub>	P <sub>Fe</sub>	P <sub>Ca</sub> % total P	P <sub>occl</sub>	P <sub>org</sub> <sup>b</sup>
6.52 +	> 2% CaCO <sub>3</sub>	16	1	< 1	62	29	7
6.03	2% CaCO <sub>3</sub>	14	3	13	24	46	15
3.91		8	6	14	28	32	20
1.99	Depth < 40 cm C < 2%, non waterlogged	56	7	19	5	41	28
1.77	Depth < 40 cm C > 2%	63	5	15	3	34	43
2.20	Depth < 40 cm C < 2% waterlogged	14	4	23	6	54	13
3.72	Fertilized	12	12	15	4	36	33

<sup>a</sup> from Ulrich and Khanna (1968).

<sup>b</sup> Number of samples: P<sub>Al</sub> = aluminium phosphates, P<sub>Fe</sub> = iron phosphates, P<sub>Ca</sub> = calcium phosphates, P<sub>occl</sub> = occluded phosphate, P<sub>org</sub> = organic phosphorus.

phosphoric acid potential ( $\text{pH} + \text{pH}_2\text{PO}_4$ ) was related either to lime potential ( $\text{pH} - \frac{1}{2}\text{pCa}$ ) or to aluminium hydroxide potential ( $\text{pH} - \frac{1}{3}\text{pAl}$ ) in the form of a solubility product diagram, the soils could be separated into four distinct groups as shown in Fig. 1. The  $\text{H}_2\text{PO}_4^-$  concentration in solution in soils with pH values > 6.8 and Ca-P as the dominant P fraction was essentially controlled by the solubility of calcium phosphates which are more soluble than hydroxy-apatite and less soluble than octacalcium phosphate. However, when soils contained more than 2%  $\text{CaCO}_3$  the concentration of P lay on or above the octacalcium phosphate line. In soils ranging in pH from 4.5–6.8, calcium phosphates changed predominantly to aluminium phosphates and the concentration of P in soil solution was not controlled by a defined phosphate form (Fig. 1). The concentration of P in solution in soils with a pH range of 3.5–4.5 was controlled by aluminium phosphates, which are more soluble than crystalline variscite, but less soluble than amorphous aluminium phosphates. This is also true for strengite and other iron phosphates for soils at pH values < 3.5. In soils with an Al-buffering system, both Al- and Fe-phosphates could occur simultaneously, the former dominating near pH 5.0 and the latter under lower pH conditions. If an acid soil is limed to increase the pH (to < 8.0) the distribution and solubility of soil P fractions do not alter.

Although the information obtained from the solubility product diagrams and the chemical fractionation method helps in postulating the possible form

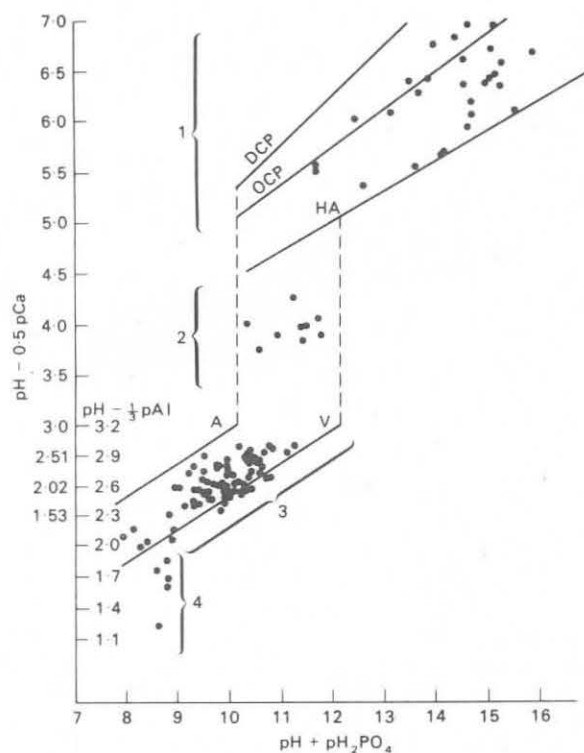


Fig. 1. The relation between the  $\text{H}_2\text{PO}_4^-$  potential and the lime or aluminium potentials for 230 German forest soils (from Ulrich and Khanna, 1968). DCP, OCP and HA are solubility lines for dicalcium phosphate, octacalcium phosphate and hydroxy-apatite respectively. A and V are solubility lines for amorphous aluminium phosphate and variscite. 1 = Range of calcium phosphate; 2 = transition from calcium to aluminium phosphate; 3 = range of aluminium phosphate; 4 = transition from aluminium to ferric phosphates.

of a compound involved in controlling the concentration of a given ion in the soil solution, knowledge of the so called Quantity/Intensity (Q/I) relationships is essential if the changes that occur in the solution and the solid phases of the soil (see Barrow, 1978), as a consequence of management practices on plantations including P fertilization, are to be understood.

The non-occluded inorganic P forms are considered important for the nutrition of plants. Adams and Walker (1975) observed a significant relationship between the amount of non-occluded soil inorganic P and the foliar concentration of P in the second rotation *P. radiata* trees growing on acid soils (yellow-grey to yellow-brown earths) of Tasman forest, New Zealand. In these soils the amount of non-occluded P ranged from 13 to

$34 \mu\text{g P g}^{-1}$  soil compared with 81 to  $185 \mu\text{g}$  total  $\text{P g}^{-1}$  soil. Fertilization with P increases the amount of non-occluded inorganic P in soils. Gentle *et al.* (1965) reported that  $96 \text{ kg P ha}^{-1}$  applied to 8-year-old *P. radiata* on a yellow podzolic soil resulted in an increase of non-occluded forms in the surface 7.5 cm. This increase was still evident 15 years after fertilization ( $56 \mu\text{g P g}^{-1}$  soil on control compared with 100 and  $93 \mu\text{g}$  for applications of superphosphate and rock phosphate respectively).

The availability to the soil solution of P from different inorganic forms depends upon the kinetics of the dissolution processes (Ulrich *et al.*, 1973) which is primarily determined by the amount of phosphorus on the surface of the components of each P form (i.e. isotopically exchangeable fractions). Ulrich and Khanna (1968) provided values for isotopically exchangeable P for a number of forest soils. When the amount of Al- and Fe-phosphates was low in soils, the fraction of them occurring as the isotopically exchangeable fraction tended to be quite high (60–80% for Al-P and 10–30% for Fe-P) indicating a low degree of crystallinity of these compounds under such soil conditions. Only under very acid soil conditions did these P forms show a high order of crystallinity, as evidenced by their low (<10%) isotopic exchangeability. Most of the phosphorous in this case is poorly available to plants.

Most of the studies on plantation soils have used the usual soil testing methods and extractives developed for agricultural systems (e.g. Ballard, 1980). Such methods lack the information required to apply the above-mentioned concept of using soil P pools, including their isotopically exchangeable fractions, for assessing the nutrient-supplying capacity of soils. Adoption of suitable methods to analyse soils on the above-mentioned concept will be an advantage in predicting P-supplying capacity of soils to plantations and in evaluating the needs for, and effects of, applying P fertilizer. Soil P forms have also been used to ascertain the residual effects of fertilizer applied to plantations (Humphreys and Pritchett, 1971).

#### (2) Organic P and its availability to plants

The role of organic P in the P nutrition of agricultural plants has recently been reviewed (Anderson 1980). Most of the organic P in plants and soil micro-organisms occurs as nucleic acids, nucleotides, sugar phosphates, phospholipids and sometimes (e.g. in seeds) as inositol phosphates. In soil, however, up to 50% of the total organic P ( $\text{P}_{\text{org}}$ ) and about 25% of the total P occur as inositol phosphates. A small fraction of  $\text{P}_{\text{org}}$  occurs as derivatives of nucleic acids (RNA, DNA) and phospholipids. About half of the organic P in soils has not yet been identified.

Inositol phosphates (e.g. myo-inositol hexaphosphoric acid or phytic acid)

are very resistant to decomposition by microbes especially when they occur in combination with a high clay content (due to adsorption, Williams and Saunders, 1956), with special clay minerals (e.g. allophane in New Zealand soils, Jackman, 1964), with high amounts of amorphous Al- and Fe-oxides (Williams *et al.*, 1958) and in freely-drained soils (Williams and Saunders, 1956). They form insoluble salts with Fe and Al in acid soils and with Ca in alkaline soils. The type of the associated cation determines their sensitivity to enzymatic attack; under acid conditions they are more resistant to enzymatic breakdown than under alkaline soil conditions.

Because little is known of the amount and rates of mineralization of  $P_{org}$ , its importance as a source of P in plantation forestry is unknown. However the importance of  $P_{org}$  for grasslands has been recognized as primarily due to the dynamics of microbial phosphorus (Cole *et al.*, 1977) which is considered to be the active fraction of  $P_{org}$ . Under *Bouteloua gracilis* sod, the cumulative mineralization during one season ( $30 \text{ kg P ha}^{-1}$ ) was closely balanced by combined uptake by plant roots and microbial population. With the methods which are now available to determine the microbial P fraction in soils (Brookes *et al.*, 1982) it may be possible to determine the significance of  $P_{org}$  for the nutrition of plantation trees. The rates of mineralization of labile organic P in 50 woodland soils have recently been assessed by using  $[^{32}\text{P}] \text{ RNA}$  (Harrison, 1982a). The rate was primarily related to soil pH and exchangeable Ca, and varied with underlying geology and season of the year (Harrison, 1982b).  $P_{org}$  may be mineralized to a greater extent under tropical than under temperate conditions. The burning of forest residues may convert part of it into inorganic form.

#### D. Sulphur Pools in Soils

Sulphur is present in forest soils in inorganic and organic forms, the latter being the predominant form in surface horizons. Organic S ( $S_{org}$ ) usually decreases with soil depth in the same way as organic C and organic N. In 19 representative forest soil profiles of northwest Germany, surface horizons (0–10 cm, A horizons) contained  $47\text{--}639 \mu\text{g total S g}^{-1}$  soil whereas deeper horizons varied from  $20\text{--}386 \mu\text{g S g}^{-1}$  (sulphate, 50–70%) (Meiwes *et al.*, 1980).  $S_{org}$  was positively related to the N content of soils especially for organic-rich horizons ( $\text{C} > 2\%$ ).

$S_{org}$ , the chemical nature of which is not precisely known, can be fractionated into C-bonded S (non-reducible with hydriodic acid, HI) and ester sulphates (reducible with HI), the latter form constituting 30–70% of  $S_{org}$  (e.g. Williams, 1975). Much of the HI-reducible S is associated with the

high molecular weight (humic acid) fraction of organic matter (Freney *et al.*, 1969). McLaren and Swift (1977) considered that the HI-reducible fraction was of greater importance than other fractions in the short-term mineralization of S, whereas C-bonded S of the humic acid fraction was mineralized in the long term, e.g. during cultivation, but it should pass through an HI-reducible form (ester sulphates) prior to its release as inorganic S.

A fraction which can be separated from the C-bonded S by the Raney-Nickel reduction consists mainly of the amino acid S (Freney *et al.*, 1975) and may account for up to 60% of  $S_{org}$  in some soils (Lowe 1965; Neptune *et al.*, 1975). The C-bonded S which is not reducible by Raney-Nickel (which may account for up to 40% of  $S_{org}$ ) because of its resistance to degradation by drastic chemical treatments has been considered to be of little value as a source of mineralizable S (Lowe, 1965).

$S_{org}$  must be mineralized to sulphate before it can be used by plants. The rate of net mineralization of  $S_{org}$  is usually very low under field conditions; Fraaij (1973) estimated that 2–3% of the total sulphur in beech and spruce litters was mineralizable.

Sulphur is added to plantation forest soils in inorganic forms (e.g. rain, absorbed gases, aerosols, fertilizers) and in organic forms, (e.g. plant litter, sludge). In plant tissues, S is present in inorganic (sulphate) and organic forms (e.g. amino acids, ester sulphates, vitamins, enzymes). The ratio of  $N_{org}$  to  $S_{org}$  in plants appears to be genetically fixed and lies between 15–17/1. In the foliage of *Pinus radiata* the N/S ratio of protein is about 16/1 (Kelly and Lambert, 1972). This ratio reduces to about 9/1 in the soil organic matter as a consequence of relative enrichment in  $S_{org}$  which implies that  $N_{org}$  must often be mineralized faster than  $S_{org}$  in the early stages of plant decomposition. The low availability of S in soils due to small inputs and due to the low amounts in parent materials such as tertiary basalt, basic extrusives and mixed granites may lead to high N/S ratios and a reduction in the sulphate content of *P. radiata* foliage if N availability is high (Lambert and Turner, 1977). Similar reduction in the amount of sulphate in the foliage of *Pseudotsuga menziesii* (Douglas fir) on application of N has been reported by Turner *et al.* (1980). Whether such reduction in sulphate leads to S-deficiency in plantation trees needs further research.

#### E. Cation Pools in Soils

A significant portion of the cations present in plantation forest soils (K, Ca, Mg, Mn, Na) may occur as structural parts of primary and secondary

minerals, or in organically bound form, and for K, fixed in the lattices of 2:1 type clay minerals. Only under special circumstances (e.g. in glacial outwash soils, Leaf, 1958) do these forms have any significance for nutrition of trees (Stone and Leaf, 1967). The soluble and exchangeable forms are thus primarily involved in the nutrition of trees.

Calcium is usually the most common cation, occupying more than 80% of exchange sites when soil pH values are over 6. As the pH decreases, the effective cation exchange capacity,  $CEC_e$ , (the sum of the exchangeable cations at the soil pH) decreases, and exchangeable Al replaces exchangeable Ca (Ulrich, 1966). This implies a change in the buffering system from a  $Ca/HCO_3^-$  to an Al-dominated system.

A number of workers have suggested that Ca may become deficient in acid soils under intensively-managed plantations (including those subjected to whole-tree logging). For example, Humphreys (1964) suggested that acid soils containing less than 1 meq exchangeable Ca  $100\text{ g}^{-1}$  soil could cause Ca-deficiency in *Pinus radiata* plantations in NSW (Australia). This deficiency is manifest in Al toxicity and P deficiency. As a plantation develops, Ca and Mg are progressively removed from the nutrient cycle as they are immobilized in the tree biomass and the litter layer (Silkworth and Grigal, 1982). However Ca deficiency is usually coupled with Al toxicity as shown in Fig. 2. A reduction in the Ca/Al ratio in the soil solution inhibits the growth of fine roots. The effect of the ratio of Ca/Al may be complicated by the concentration of P in the system, so that interactions among Ca, Al and P are strongly involved.

Factors affecting the availability of K in soils have been recently reviewed by van Diest (1978). Potassium deficiency may particularly occur in plantations on sandy and calcareous soils. Stone (1958) observed K deficiency in pines when the amount of this cation was less than 0.05 meq  $100\text{ g}^{-1}$  soil. Potassium applied to plantations in a single application can persist over a long period of time because it undergoes a rapid turnover in the system. With the help of the "reverse tracer technique", which involves determination of relative proportions of Rb and K in the foliage, from fertilized plots Stone (1981) observed that 25–45% of K in the foliage, of a number of the tree species examined, had its origin in K-fertilizer which was applied 4–5 years earlier.

#### F. Micronutrient Pools in Soils

Micronutrients in forest soils may be present in many forms. Nevertheless the information provided by Stone (1968) in his review on the microelement



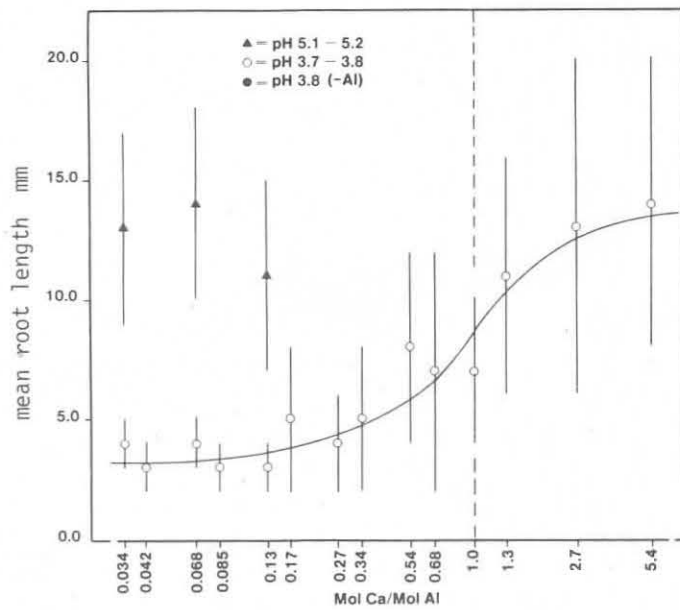


Fig. 2. Growth of the roots of *Picea abies* seedlings as a function of Ca/Al ratio in the solution. The closed circle represents the value for solutions adjusted at pH 3.8 but without Al. The open circles represent solution where Al is present as  $Al^{3+}$  (pH 3.7 to 3.8). The closed triangles indicate values at pH 5.1 to 5.2, where Al occurs as Al-hydroxocomplex and is non-toxic to roots (from Rost, 1982).

nutrition of forest trees can be used to identify soils on which, plantation forests may commonly show micronutrient deficiencies (micronutrients involved are in brackets):

- (1) Soils derived from deficient parent materials. Soils involved are mostly sandy and lack in primary minerals (B, Cu, Fe, Mn, Zn);
- (2) acid soils subject to leaching (B, Cu, Mo);
- (3) acid peat and muck soils (B, Cu, Mo);
- (4) calcareous soils containing free lime in the fine earth fraction and heavily limed acid soils (B, Cu, Fe, Mn, Zn);
- (5) soils containing excessive quantities of P (Cu, Zn), Mn (Fe) and N (Cu);
- (6) gley or pseudogley soils which are low in soluble Fe or high in soluble Mn (Fe);
- (7) serpentine soils, soils containing 'ironstone' and laterites (Mo).

Raupach and Clarke (1978) observed that soils with high contents of goethite, interstratified 2:1 lattice silicates and organic matter may limit the

availability of Cu, Zn and Mn in *P. radiata* stands, especially under swampy conditions in podzols. They suggested the use of soil survey of the main soil types together with an assessment of tree size and foliar nutrient status within each soil group may be a useful approach to explore the deficiencies and toxicities of elements.

### III. NUTRIENT TRANSFERS IN SOILS

Nutrients in forest ecosystems are transported chiefly by water and in the biomass, though nutrients such as N, S, P and B may also be transferred in gaseous form. Transfer in gaseous form usually involves a number of oxidation/reduction processes, such as the production of  $N_2$ , NO,  $N_2O$  from organic N, the production of  $H_2S$  and  $NH_3$  from organic S and N respectively and the production of  $N_2$  and  $N_2O$  from  $NO_3$ . Many of these processes are brought about by microbial activity. Although of importance for plantation soils, they will not be examined in this chapter.

Nutrient transfer processes which are mediated by organic matter in a forest ecosystem include the production and decomposition of litter, accumulation by vegetation and removal of biomass. Some of these processes are discussed in Chapters 3 and 5. The discussion here will be restricted to the process of nutrient transfer in soils through the soil solution.

Most management practices in a plantation will have some effect on the chemical composition of the soil solution. The important ones include the operations during site preparation (cultivation, fertilization and weed control), silvicultural treatments applied during the stand development (e.g. pruning, thinning, fertilization, prescribed burning, irrigation, the removal of canopy cover), and the operations which take place during harvesting (use of heavy machinery causing soil compaction and erosion, and slash burning). Some of the changes in the chemical composition of soil solution are the direct consequence of adding nutrients to the soil, whereas others arise indirectly through processes such as changing the rates of nutrient uptake, mineralization and immobilization, and exchange and weathering processes.

#### A. The Chemical Composition of Soil Solution and the Uptake of Ions by Roots

The composition of the soil solution and the uptake of nutrients by roots vary spatially (both on a macro and a micro scale) with the season, and as a result of management practices. Moreover, the uptake of nutrients depends

upon the absorption capacity of roots as well as on the mass flow and diffusion of ions to root surfaces (see Chapter 6). Consequently it is not feasible to predict either total annual uptake, or the seasonal pattern of uptake, using only information on the chemical composition of the soil solution. Total uptake can be calculated from other independent parameters such as rates of elemental retention in the biomass, and other elemental fluxes such as occur in litterfall and throughfall (see Chapter 3). It is possible to assess, however, the contribution that mass flow makes to total element uptake by comparing total uptake, water lost by transpiration and nutrient concentration in soil solution. The fraction of ion uptake resulting from mass flow has been defined as the mass flow coefficient (MFC) by Prenzel (1979). He calculated coefficients for 11 elements in the soil of a 125-year beech forest (on an acid brown earth) in northwest Germany. The mass flow coefficients of the elements could be grouped into three classes (see below): (a) those (Al, Cl, Na) with values  $< 1$ , which are selectively discriminated against during uptake; (b) those (Mn, Ca, K, N, P) with values  $> 1$ , which are selectively taken up; and (c) those (S, Fe, Mg) with values  $\sim 1$ , for which mass flow accounted for almost the same amount as was actually taken up by the stand:

Element:	Al	Cl	Na	S	Fe	Mg	Mn	Ca	K	N	P
MFC:	.076	.086	.34	.77	1.5	1.7	2.2	2.2	8.3	11	120
Major mechanism of uptake:	Discrimination			Mass flow			Selective uptake				

Such studies, which provide insight into the relative importance of various physiological processes in ion uptake (see Chapter 6) may help to explain different rates of tree growth under a range of soil conditions.

#### B. Mineralization Processes and the Chemical Composition of Litter Leachates

Either net mineralization or net immobilization of nutrients accompanies the decomposition of litter, thereby affecting the composition of the soil solution. A comparison between the composition of the solution leaving the humus layer with that entering it (e.g. from throughfall) thus provides an indication of the degree and rate of decomposition. Mayer (1972), assuming a steady-state condition for the litter layer under a beech stand (soil type: acid brownearth, humus type: moder), used the following equation to describe the rate of decomposition of humus in the absence of uptake:

$$N_{(out)} = N_{(in)} + X_{(t)} - X_{(t-1)} + D$$

where  $N_{(out)}$  is the amount of a nutrient collected below the humus layer,  $N_{(in)}$  the amount entering the humus layer via throughfall,  $X_{(t)} - X_{(t-1)}$  represents the amount of nutrient released through mineralization during the period between  $t$  and  $t-1$  months which can be obtained from an exponential function of the type ( $X = X_0(1 - e^{-kt})$ ), and  $D$  is a function of rainfall and temperature. He observed the following trends in the release of nutrients from a humus layer:

- (a) The amount of nutrient released varied in the order  $K > P > N > Ca$ ,  $Mg$ . A similar order has been observed for oak stands by Remezov and Pogrebnyak (1969).
- (b) The amounts released were higher during autumn than during the rest of the year. This was after leaf fall and was probably due to leaching out of leaves and microbial catabolism of less resistant substrates. A similar result has been reported by Feller (1978) under *Pinus radiata* in Victoria. High values have sometimes been reported in spring (Grier and Cole, 1972). McColl (1972) observed that the concentrations of ions in forest-floor leachates under *Pseudotsuga menziesii* increased as air temperature increased, length of the drying period between storm fronts increased, and the quantity of precipitation decreased. Concentrations of elements in litter leachates in various forest ecosystems may range from 0.8–3.9  $mg\ l^{-1}$  for  $Na$ , 1.7–7.2 for  $K$ , 2.0–6.6 for  $Ca$ , 0.7–1.8 for  $Mg$ , 4.2–7.9 for  $Cl$  and 4.5–16.3 for  $SO_4-S$  (Raison and Khanna, 1982), and usually follow the sequence: litter leachates  $>$  throughfall and soil solution.

The pH of leaves of temperate deciduous trees normally lies within the range of pH 5.0–6.5 while that of conifer needles shows values of 3.5–4.2 (Handley, 1954). Because the pH of leaf litter usually tends to follow the base content of soils, the pH of hardwood and conifer litter may be closely coupled to the soil types on which the trees are growing. Additionally, plants growing on base-deficient soils allocate a greater proportion of net primary productivity to plant protection chemicals such as phenolic acids, and thus increase the acidity of the soil (Swift *et al.*, 1979).

### C. Chemical Composition of the Soil Solution and the Leaching of Nutrients

Most studies of the chemistry of soil solutions under plantations have sought to quantify the extent and the rate of leaching of nutrients out of the rooting zone, and to define the factors responsible. Cations and anions are transported together, by percolating solutions, and because the capacity of the

soils to retain anions is generally low, either the external input of anions, or their production within the soil in amounts exceeding the uptake capacity of root, will tend to increase the leaching of nutrients. The factors responsible for the presence of excessive anions in plantation soils vary. For example, an excess may be due to excess  $\text{Cl}^-$  in areas influenced by high rates of accession of marine salts, excess  $\text{SO}_4^{2-}$  or  $\text{NO}_3^-$  from man-made atmospheric emissions,  $\text{HCO}_3^-$  from microbial and root respiration in soils,  $\text{NO}_3^-$  added as fertilizer or produced during mineralization of organic N in soils, and the release of organic anions from decomposing organic matter. Forest management practices such as harvesting (clearcutting), and the addition of soluble fertilizers and cultivation, which lead to excessive production of anions in the system, will consequently increase the leaching of nutrients. The consequences of such practices have been discussed in Chapter 14.

The nature of the cations which accompany the excess anions in the leaching process is controlled by the establishment of equilibria between soluble and exchangeable phases. The extent and direction of such exchange reactions is determined by the gradient in the Schofield's potentials (SP) which are analogous to the potential gradients in soil water (Ulrich, 1969). Gradients in SP (Table II) provide a measure of the reaction potential with which an ion is released or retained by the soil components, and thus enable predictions to be made about the sinks and sources of ions at different soil

Table II. Gradients of Schofield's chemical potentials between horizons for cations during vertical transport of soil solutions in a *Picea abies* plantation.<sup>a</sup>

Transport From/To	Chemical potential ( $\text{kJ mol}^{-1}$ )					
	H	Na	K	Mg	Ca	Al
0-horizon						
0-10 cm	-6	-9	+8	+2	+5	-1
0-10 cm						
10-20 cm	+9	-3	+8	-5	-3	-6
10-20 cm						
20-30 cm	+2	0	-1	—	+5	-1
20-30 cm						
30-40 cm	+2	-2	0	+1	+1	+5
30-40 cm						
40-50 cm	+5	-1	0	-4	0	+1

<sup>a</sup>From Ulrich *et al.* (1979); No. 5 in Table IV.

depths. Table II indicates a source of  $H^+$  ions in the solution as it infiltrates from the humus layer into the 0–10 cm layer. Ulrich *et al.* (1979) ascribed this source of  $H^+$  ions to the formation of organic complexes with Al and Fe. Below the 10 cm depth the soil acts as a sink for  $H^+$  ions. There is no tendency for  $Na^+$  to be retained by the upper 50 cm of soil. The sink for  $K^+$  is confined to the upper 20 cm depth, most probably due to its uptake by roots from these soil layers. The release of  $Al^{3+}$  from soil components occurs mostly in the 0–10 cm horizon.

These observations on the behaviour of ions in the percolating solution are valid only for equilibrium conditions. Under field conditions the attainment of equilibrium between ions in the percolating soil solution and the soil is usually slow and may not always be fully achieved due to soil and plant processes, including mineralization, mobilization and nutrient uptake.

#### D. Mineral Weathering

The contribution of mineral weathering under plantation management to the supply of plant nutrients to trees is controlled mainly by the processes of chemical equilibria in soils. Methods of assessing weathering rates include: (1) mass balance techniques which can be applied to either watershed studies (Likens *et al.*, 1977) or to lysimeter studies (Ulrich *et al.*, 1979); (2) descriptive field studies of weathering for quaternary stratigraphic research (Birkeland, 1974); (3) laboratory studies of nutrient release from soil minerals (Thompson *et al.*, 1977); and (4) the use of thermodynamic relationships to predict soil mineral transformations (Kittrick, 1977). All have limitations in predicting the rates of release of available nutrients in the rooting zone of trees.

Using mass balance techniques, Clayton (1979) derived values for the rates of release of elements from rock weathering in the USA which ranged from traces to 1.2, traces to 0.3, traces to 4.3 and 0.2–4.3  $keq\ ha^{-1}\ yr^{-1}$  for Na, K, Ca and Mg respectively. Johnson *et al.* (1981) calculated from the ionic denudation rate in the streamwater and by biomass retention that 1.8–2.0  $keq\ ha^{-1}$  of elements are released annually by chemical weathering at the Hubbard Brook Experimental Forest, New Hampshire, USA. Ulrich *et al.* (1979) obtained values for the rate of release of elements from sedimentary parent materials from the consumption of  $H^+$  ions during the weathering of silicates. For the rooting zone (50 cm) they found that  $H^+$  consumption was about 0.2  $keq\ ha^{-1}\ yr^{-1}$  under a beech stand at Solling in West Germany. This was equivalent to an annual release of 1.6  $kg\ Mg^{2+}\ ha^{-1}$ , probably from biotite minerals. An estimation of the release of nutrients on mineral

weathering can vary manifold (tenfold for the two examples given above). The contribution of released nutrients towards the annual uptake values (see Table IV), though significant in some cases, remains therefore speculative.

#### IV. SOIL CHEMICAL CHARACTERISTICS AND PROTON BUFFER SYSTEMS

Forest ecosystems are dynamic; the driving force for changes in the chemical status of soil components of a forest ecosystem are protons or the acids which generate them. The nature of the proton buffer systems determines the chemical status of the soil and thus the environment for decomposer organisms and roots. Soil processes such as the accumulation and mineralization of organic matter, the weathering of rock minerals, the uptake of ions by roots and the leaching of elements out of the rooting zone involve the production or consumption of hydrogen ( $H^+$ ) ions. Between tree rotations, a balance between ion uptake and mineralization of organic matter is of prime importance. In many forests, particularly conifer plantations, organic matter (litter) accumulates on top of the mineral soil. The rate of accumulation is especially high after closure of the forest canopy. This means the uptake of ions is greater than that released via mineralization processes in the ecosystems until a steady-state occurs. Such a decoupling of ion uptake and release has two consequences:

- (a) production of protons (acids) within the soil which promotes podzolization (Ulrich 1981a); and
- (b) withdrawal of nutrients, especially N and P, from the nutrient cycle, thus diminishing the easily-mobilizable pools.

Both effects can result in the development of nutrient deficiencies. Acidification may lead to leaching of micronutrients such as Mg, Mn, Co, Zn, Cu and B from soil to the extent that the supplies are inadequate for optimum tree growth. Accumulation of N, P, Mg and Ca in undecomposed litter residues can accentuate existing inadequacies. The nutrient cycle may be a closed one at the time of harvesting and planting, depending primarily on how the plantations are managed during those phases, i.e. primarily on the rate of decomposition of organic matter and the losses of nutrients associated with it. If management practices, including the introduction of exotic tree species, lead to soil acidification, the rotational cycle may no longer be closed and the ecosystem then undergoes long-term unidirectional development in the form of a succession if left to go unchecked. In some cases

the end-point of this succession will be the development of a savannah or heath ecosystem. To reach the end-point of such a succession would probably need several tree rotations of 40–100 years each, although the process may be accelerated by the intensive utilization of biomass e.g. litter raking or whole-tree harvesting (Ulrich, 1981b). The consequences of such practices may be minimized by balanced fertilization. The natural processes counteracting soil acidification are the dissolution of  $\text{CaCO}_3$  and the formation of silicic acid from silicates (weathering). Both of the resulting acids ( $\text{H}_2\text{CO}_3$  and  $\text{H}_4\text{SiO}_4$ ), are converted to oxides ( $\text{CO}_2$ ,  $\text{SiO}_2$ ) and water, which means that the protons are returned to water. The rate of  $\text{H}_4\text{SiO}_4$  liberation from silicates will, in many soils, be high enough to counteract the proton production caused by the removal of stem wood in harvesting (Ulrich *et al.*, 1979). In most cases, however, it will not be sufficient to counteract the proton production caused by intensive utilization of biomass. It is obvious that acid rain enhances this unidirectional development of the forest ecosystem.

In order to assess the significance of the processes described above, forest soils can be grouped into five buffer ranges based upon the stability of soil minerals and oxides, and the possible buffering reactions involving protons (Ulrich, 1980). For purposes of simplification, only well aerated and freely draining soils are considered:

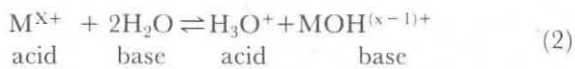
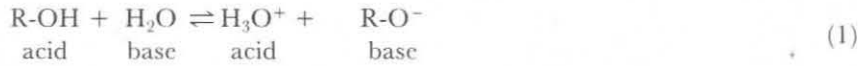
- (a) Calcium carbonate buffer range (pH > 8–6.2).
- (b) Silicate buffer range (pH 6.2–5.0).
- (c) Cation exchange buffer range (pH 5.0–4.2).
- (d) Aluminium buffer range (pH 4.2–2.8).
- (e) Iron buffer range (pH 3.8–2.4).

These pH values are those in the soil solution, although they are commonly described as soil pH values. In most soils they can be satisfactorily obtained by the pH measurement of the equilibrium soil solution of a soil/water suspension [ $\text{pH}(\text{H}_2\text{O})$ ]. If the pH drops after the addition of KCl to the soil/water suspension, the lower values indicate the buffer range which is attained during periods of high proton or salt load (e.g. following fertilization).

Intensity (pH values) as well as capacity parameters are needed to describe soil acidity. The pH is a measure of the mean free energy per proton, i.e. it gives the mean acid strength of the dissociating groups. The capacity term is the base neutralization capacity which is equivalent to the sum of all acids which could be neutralized with a strong base up to pH 7. The capacity term in a soil is difficult to determine because it may include some very slow reactions. In soils, two different kinds of Brønsted acids,  $\text{R-OH}$  and  $\text{M}^{\text{X}+}$ ,



can be distinguished showing the following proton transfer reactions:



It is an important feature of soil chemistry that  $\text{MOH}^{(x-1)+}$  can react as a base as well as an acid:



In these reactions R symbolizes carbon groups (mainly phenolic hydroxyl groups and carboxyl groups), and Si (associated with pH-dependent charges of clay minerals, and dissolved silicic acid at  $\text{pH} > 7$ ); M refers to metal cations like Al, Mn, Fe and heavy metals including their anionic complexes. An accumulation of the positively charged ions of these metals in acid soils would therefore mean a potential for proton transfer.

#### A. Calcium Carbonate Buffer Range

Only those soil horizons which contain  $\text{CaCO}_3$  in their fine-earth fraction stay exclusively in this buffer range. In these horizons the dissolution rate of  $\text{CaCO}_3$  is high enough to keep the system close to the equilibrium described:



The common acid is carbonic acid resulting from respiration by roots and microbes decomposing organic matter. The buffering capacity amounts to  $150 \text{ kmol H}^+$  per 1%  $\text{CaCO}_3$ . The buffer rate is high as long as it is determined by the rate of dissolution of  $\text{CaCO}_3$ , and is not limited by diffusion due to uneven distribution of  $\text{CaCO}_3$ .

Ca is the predominant cation in the soil solution and on the exchange complex of calcareous soils. Examples of such soils (calcareous brown soils and sierozems) containing 0.4–46.3%  $\text{CaCO}_3$  and carrying *Quercus* plantations have recently been described by Rafahi (1982). Depending upon the rate of  $\text{CO}_2$  production and the soil pH, the concentration of  $\text{Ca}^{2+}$  and  $\text{HCO}_3^-$  ions in the soil solution may be high. Such soils have a stable structure. Calcium phosphate is the main form of inorganic P. The absence of toxins such as  $\text{Al}^{3+}$  ions and water-soluble phenols means that there is no hindrance to the growth of plant roots and the activity of bacteria with the

result that there is a rapid turnover of litter (thus producing mull-type humus) if temperature and moisture of the soil are favourable. The soil organic matter formed has a high N content and a low C/N ratio, approximately 10. Nutritional factors which would limit the growth of forest trees in such soils are:

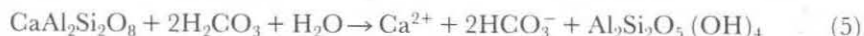
- (a) Unfavourable cation ratios due to a high concentration of  $\text{Ca}^{2+}$  at the expense of other cations. A low  $\text{K}^+/\text{Ca}^{2+}$  ratio could lead to plant deficiency in K e.g. where the exchangeable K values in the surface calcareous soils ranged from 0.3 to 0.9 me  $100\text{ g}^{-1}$  (Rafahi, 1982).
- (b) The fixation of  $\text{K}^+$  and  $\text{NH}_4^+$  on illitic clay mineral particles which would limit their availability for plant growth.
- (c) Unbalanced cation/anion ratios in ion uptake may be ~~unbalanced~~ due to a large anion surplus (e.g. percentage of  $\text{NO}_3^-$ ) in some systems. For example, Le Tacon (1978) observed that the poor growth of *Picea abies* (Norway spruce) in calcareous soils was due to the low availability of  $\text{NH}_4^+\text{-N}$  which produced an imbalance in cation/anion uptake and a high Ca content in leaves. In contrast, the growth of *Pinus nigra* (Austrian pine) on the same soils was unaffected by the presence of calcium carbonate because the pine is able to utilize nitrate-N and to tolerate calcium carbonate through mycorrhizal associations.
- (d) A low solubility of metal trace elements such as Mn, Fe, Cu, Zn which may result in their deficiencies (see Section II, F).

The non-availability of metal trace elements in this buffer range cannot be easily corrected by adding soluble forms to the soil because these are readily transformed into compounds of low solubility. The same will be true for K and  $\text{NH}_4$  fertilizers if added to soils dominated by illitic clay minerals. It may be possible to overcome some of these problems of nutrient availability by introducing and maintaining a herb layer including legumes, which would have the effect of enhancing the rate of turnover of nutrients in the ecosystem. The high rates at which phytomass will form under such conditions will increase the storage of organic N in the soil. High turnover rates of organic matter may narrow the cation/anion ratio in ion uptake by driving the plants to take up some N as  $\text{NH}_4\text{-N}$ , and can also improve the uptake of metal trace elements as organic chelates.

#### B. Silicate Buffer Range

As the pH (in  $\text{H}_2\text{O}$ ) values of soils approach 6,  $\text{CaCO}_3$  ceases to act as a buffer and is absent at least from the fine earth fraction of these soils. As long

as carbonic acid is the only acid being produced in the soil, or involved in equilibrium reactions, the pH ( $H_2O$ ) of soils will stay at values  $> 5$ . In this range the only buffer system acting is through the weathering of silicates, which releases alkali and alkali earth cations. For example the weathering of Ca feldspar to kaolinite:



The cations liberated during the weathering of silicates are bound as exchangeable cations in the clay minerals formed from the weathered silicate lattices. The protons consumed during this process are converted to undissociated silicic acid which is finally broken down to  $SiO_2$  and  $H_2O$ . Any leaching of silicic acid is not accompanied by cations. There are low concentrations of  $HCO_3^-$  as the pH approaches 5.0, and of  $NO_3^-$  because the rate of nitrification usually does not exceed the rate of nitrate uptake. As a result, leaching losses from the soil are low, irrespective of the amount of rainfall.

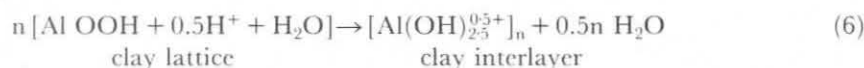
The buffer capacity of such soils as well as the buffer rate depends upon the content and type of silicates involved. On average, the buffer capacity amounts to  $25 \text{ k mol } H^+$  per 1% silicate. The buffer rate may vary between 0.2 and  $2.0 \text{ k eq ha}^{-1} \text{ yr}^{-1}$  (compare with values given in Section III, D); the values are  $< 1$  in soils formed from sedimentary rocks (Ulrich, 1980).

As with calcareous soils, the soils in this buffer range exhibit relatively small gradients in the chemical soil characteristics with depth. The soils have a crumb-like structure, the organic matter in the root zone has a low C/N ratio and the humus is of the mull-type.

A high turnover rate of nutrients and a large amount of nutrients in the rooting zone of these soils ensure a reasonably well balanced and adequate nutrient supply to plants. Forest managers should therefore aim at bringing the soils to or maintaining them in this buffer range, by liming if necessary.

#### C. Cation Exchange Buffer Range

If protons remain on the surface of clay minerals for periods of weeks or months, they are consumed and release Al ions from the lattice of clay minerals (Schwertmann and Jackson, 1964). Between pH 5.0 and 4.2, most Al ions form polymeric hydroxo cations with a charge of  $+0.5$  per Al atom (equation 6), which accumulate in the interlayers of the swelling clay minerals. They are not easily exchangeable if soil pH increases or there is a decrease in pH.



The buffering capacity is equal to the cation exchange capacity, i.e. about  $7 \text{ k mol H}^+ \% \text{ clay}^{-1} \text{ ha}^{-1} \text{ dm}^{-1}$ . The buffer rate is usually sufficiently high to prevent soils from entering the Al buffer range, i.e.  $< \text{pH } 4.2$ , as long as the degree of exchange saturation of Ca is not low,  $> 10\%$ .

The ratio of exchangeable Al to exchangeable cations in such soils tends to vary from 0.2 to 0.9, and the effective CEC/total CEC ratio from 0.1 to 0.7. The more the exchange sites are occupied by Al, the less they are available for K, Mg and Ca, which are consequently leached. These soils may have sufficient Al in soluble and exchangeable form to be toxic to calcicole species in particular, with the result that there is a change of forest flora. The biotic activity in the soil (including that of earthworms) is reduced and litter tends to accumulate on the forest floor (moder or mor humus). The development of a mor layer results in immobilization of N, P and S. Phosphates exist in bound forms associated with Fe and Al which have lower solubility than those in the silicate buffer range. These soils are therefore likely to show nutrient deficiencies, and the application of fertilizers would probably increase phytomass production and forest yield.

Many acid forest soils lie within this buffer range. They usually exhibit marked gradients in chemical characteristics with depth. Where there is disintegration of soil organic matter (see Section V, A), the lower soil horizons may be more acid than the surface layers. However, the opposite may commonly occur if the lower soil horizons are still in the silicate buffer range.

#### D. Aluminium Buffer Range

At soil pH values  $< 4.2$ ,  $\text{Al}^{3+}$  reaches concentrations in the soil solution which make it the most important factor in determining the stability of forest ecosystems. The sum of exchangeable H, Al, Mn and Fe ions (exchangeable Al being dominant), which represents the potential acidity, will exceed 90% of the  $\text{CEC}_r$  of the soil. Exchangeable Ca and Mg are reduced to values less than 5–10%.

The buffer capacity amounts to  $100\text{--}150 \text{ k mol H}^+ \% \text{ clay}^{-1}$  depending upon the type of the clay minerals. The buffer range depends upon the dissolution rates of the solid phases. It is high (several  $\text{k mol H}^+ \text{ ha}^{-1} \text{ yr}^{-1}$ ) for the intermediate type of Al compounds, e.g. polymeric Al hydroxo cations and Al hydroxo sulphates, and low ( $< 1 \text{ k mol H}^+$ ) for Al released from silicate lattices. As a consequence, a soil horizon remains in the aluminium buffer

range as long as Al hydroxo compounds with high dissolution rates are available.

The toxicity of Al is a well-known phenomenon in plants (Foy *et al.*, 1978) and has been demonstrated in seedlings of beech and pine (Süchting, 1943, 1948). However, the extent of toxicity depends upon the nature of the ionic species of Al present in the soil solution, and the ability of roots to change the form of Al in the cell walls before it reaches cell membranes. Monomeric species, which are high in concentrations at pH < 4.2, possess the highest potential toxicity, whereas polymeric Al species and organically chelated Al show the lowest potential. It is not the concentration of monomeric Al, but its ratio to Ca in the soil solution, which determines the damage to roots (Rost, 1982).

This interaction between Ca and Al is evident from the data in Table III. Data on Ca/Al ratios in the equilibrium soil solution and in fine roots come

Table III. Content of Ca and Al in fine roots from soils containing varying fractions of Ca,  $X_{Ca}^S$ , on the exchange complex, and from the litter OH-horizons.

$X_{Ca}^S$	No. of samples	Composition of fine roots					
		mol Ca <sup>b</sup> mol Al	Ash content (%)	SiO <sub>2</sub> (%)	Ca (mg g <sup>-1</sup> dry wt)	Al dry wt)	mol Ca mol Al
> 0.5	19	75	6.7	3.0	8.8	3.6	2.3
		+ 49	+ 2.7	+ 2.5	+ 2.4	+ 1.6	+ 1.5
0.5–0.1	9	15	7.3	4.2	5.2	2.3	1.6
		+ 14	+ 2.3	+ 2.4	+ 3.0	+ 0.6	+ 0.9
0.1–0.05	18	3.5	8.2	5.3	2.1	4.1	0.4
		+ 6.2	+ 3.5	+ 3.5	+ 1.4	+ 1.8	+ 0.3
< 0.05	46	0.96	8.8	5.8	1.6	4.8	0.3
		+ 0.52	+ 4.9	+ 4.0	+ 0.8	+ 2.2	+ 0.2
OH-Horizons	11	—	3.5	1.0	5.8	2.1	1.8
		—	+ 0.5	+ 0.5	+ 4.0	+ 0.6	+ 1.0

<sup>a</sup>Ca as a fraction of the total exchangeable ions.

<sup>b</sup>Mean + S.E. in the equilibrium soil solution (saturation extract).

from a variety of soils of variable exchangeable Ca percentage,  $X_{Ca}^S$  (> 50%–< 5%). Ca/Al values for roots developing in the litter layer (usually moder and raw humus) have also been included. An increase in acidity (decreasing Ca saturation) is reflected in a marked decrease in the Ca content of fine roots, but only a small increase in their Al content. However, the Ca/Al

ratios varied significantly from 2.3 at Ca saturation of  $> 50\%$  to 0.3 at Ca saturation of  $< 5\%$ . Roots from the litter layer had low Al content and a high Ca/Al ratio pointing to the decreased effects of Al toxicity due to the complexing of Al by organic matter. A Ca/Al (in moles) ratio of  $< 1$  in roots indicates Al toxicity. In addition, soils with the following characteristics will have a high potential for Al-toxicity to tree species possessing some tolerance to Al: (a) mineral soils showing less than 5% exchangeable Ca saturation; (b) top soils where the ratio of Ca to the sum of Ca, Al and Fe (all expressed in equivalent weight terms) in the organic layer is less than 0.05; and (c) soils where the molar ratio of Ca to Al in the soil solution is  $< 1$ .

Soil horizons in the aluminium buffer range usually contain small quantities of cations, organic matter and N. The solubility of soil P is usually low. On the other hand, the heavy metals, including trace elements, are readily soluble and thus susceptible to leaching losses. Mn may occur in toxic concentrations (Adams and Walker, 1975). Consequently many factors limit the growth of vegetation on these soils.

The application of water-soluble fertilizers containing nitrates, chlorides and sulphates on these soils would increase Al toxicity. A number of cases are known where detrimental effects of fertilization have been observed (Waring 1973), but these are not usually ascribed to Al toxicity.

#### E. Iron Buffer Range

At soil pH values  $< 3.2$  (but in horizons influenced by infiltrating organic matter already at  $\text{pH} < 3.8$ ), the effects of Fe ions become important, and follow the same principles as described for soils of the Al buffer range. The main difference, however, is that the transport of Fe leads to visible colour changes in the soil profile (podzolization) whereas the transport of Al does not. Such soils, which contain more toxins and show greater nutrient deficiencies, have greater adverse effects on growth than those controlled by the Al buffer range. However, the process of podzolization may occur in phases of only limited duration, and the presence of a eluvial (bleached) horizon may not necessarily mean that the process is currently active.

### V. ELEMENTAL CYCLES AND PROTON BALANCES IN FOREST ECOSYSTEMS

All ecosystem processes relating to ion turnover can be reduced to four major nutrient fluxes: input (e.g. from atmosphere or by weathering of soil

minerals), output (e.g. leaching, denitrification), uptake from soil solution into biomass (by plants and micro-organisms), and liberation of ions during the decomposition of organic matter (mineralization) (see Chapter 3). For an ecosystem where the input and output fluxes for all elements are equal, the system could be assumed to be in a steady state if no changes of a temporal or a spatial nature occur in the total reserves of nutrients and biomass and in their binding forms. Under conditions approaching a steady state, the amount of nutrients taken up by plants for production of biomass is balanced by the amount mineralized from organic sources. The effects of perturbations (due to management practices) on site are discussed in Chapter 14.

Because the processes of mineralization and uptake of ions are associated with the consumption and production of protons, they result in pH changes in the soil. So far these processes have been studied in isolation, e.g. the uptake of N as  $\text{NH}_4^+$  or  $\text{NO}_3^-$ . The effects of such processes on ecosystems can, however, be judged only if all ions are considered so that the principle of electroneutrality applies in all the compartments of a forest ecosystem.

Among the elements taken up by forest trees, N is often dominant (Table IV). The total uptake rate is calculated as the sum of the rates of nutrient transport in litter (roots included) and of canopy leaching, and of accumula-

Table IV. Annual uptake of elements by forest ecosystems.

No.	Vegetation	Age (yr)	Soil type	Element (k mol ion eq ha <sup>-1</sup> )					
				N	P	K	$\frac{1}{2}\text{Ca}$	Cations (excluding N)	Anions
(1)	<i>Pinus radiata</i>	4-8		7.6	.45	2.0	.99	3.67	—
(2)	<i>Pinus patula</i>	18-20	Latosol and Andosol	10.4	.4	2.0	5.4	—	—
(3)	<i>Cupressus lusitanica</i>		Latosol and Andosol	7.3	.3	1.9	8.9	—	—
(4)	<i>Fagus silvatica</i>	120	Podzolic acid brownearth	5.8	.27	1.2	1.9	5.3	1.01
(5)	<i>Picea abies</i>	90	Podzolic acid brownearth	3.6	.14	.5	1.5	2.8	.80
(6)	Temperate deciduous forests	—	—	6.3	.22	1.1	2.1	4.6	—
(7)	Temperate conifer forests	—	—	2.9	.15	.5	.9	—	—

(1) Madgwick *et al.* (1977), gross uptake values by above-ground vegetation; (2) and (3) Lundgren (1978), above-ground vegetation; (4) Ulrich (1981), includes roots; (5) Ulrich *et al.* (1979), above-ground vegetation, (6) and (7) Khanna (1974), above-ground ~~nutrients~~ <sup>vegetation</sup>.

tion in biomass. In Table IV data are given for a 120-year *Fagus sylvatica* stand and for a 90-year *Picea abies* stand growing on a podzolic brown earth. As N may be absorbed as  $\text{NH}_4^+$  or  $\text{NO}_3^-$ , the cation/anion balance (salt balance) resulting from the uptake of nutrients is markedly affected by the form in which N is taken up. The form of N uptake is often unknown, but would depend upon soil conditions as well as on the nitrate reductase activities of root tissue—a species-determined characteristic (Clement, 1977). Assuming that uptake of N occurs solely as  $\text{NO}_3^-$ , an anion excess of 1.5 k mol ion eq  $\text{ha}^{-1} \text{yr}^{-1}$  (1.0 + 5.8–5.3) for *F. sylvatica* and 1.6 k mol for *P. abies* will result (Table IV), which will cause an increase in the pH value of the rhizosphere soil. However, if the uptake of N occurs solely as  $\text{NH}_4^+$ , an excess of cations of 10.0 k mol for *F. sylvatica* and 5.6 k mol for *P. abies* (2.8 + 3.6–0.8) would produce an equivalent amount of  $\text{H}^+$  ions in the rhizosphere. Values of this order are significant when they apply for long periods of time.

transfers would be compensated by those associated with the mineralization processes. However, most ecosystems are not in a steady state, though they may tend towards it, which means that phytomass production and mineralization do not compensate each other for the production and consumption of protons in soils.

#### A. Acidification and Deacidification During Forest Development

After canopy closure, organic matter on the forest floor often continues to accumulate until the stand reaches "pole size". From this stage until maturity the system normally undergoes cycles of humus accumulation and decomposition associated with thinning operations. The export of biomass and the removal of slash during clearfelling create conditions in which the decomposition of organic matter is enhanced, leading to the formation of nitrates (Ulrich and Wachter, 1971). The phase of humus disintegration is not confined to surface soil layers, but also occurs in deeper soil horizons, and remains active for a great proportion of the rotation. A number of physical and chemical changes in soils (decrease in pH and organic matter, increase in bulk density, mineralization and leaching of N and toxicity by Al etc.) which have been associated with this phase are considered to be at least partly responsible for the depressions in yields which occur in shifting agricultural systems, and in second and succeeding rotations of plantations (Ulrich, 1980). The proton balance in this phase is usually positive, leading to acidification as shown by the following calculation:



Let us assume that 100 tons of soil organic matter are mineralized in 50 years and that its C content is 50%, C/N ratio is 10 and CEC is  $4 \text{ eq kg C}^{-1}$ . If N were completely nitrified (no denitrification) it would result in the consumption of  $200 \text{ k mol H}^+ \text{ ha}^{-1}$  during mineralization and in the production of  $360 \text{ k mol H}^+ \text{ ha}^{-1}$  during nitrification. The net annual proton production would amount to  $3.2 \text{ k mol H}^+ \text{ ha}^{-1}$ . This value is generally higher than that which can be compensated for by silicate weathering. If  $\text{NO}_3^-$  is lost by leaching from the system after clearcutting and cultivation, there will be an additional load of  $\text{H}^+$  on the system, which occurs with a greater intensity (i.e. high value for a short period) (Ulrich *et al.*, 1979). Net losses (output-input) of  $\text{NO}_3^-$  and basic cations from the rooting zone are the best measures of humus disintegration.

During the humus aggradation (build up) phase, protons will be consumed from the soil system. In a case similar to that above, about  $3.2 \text{ k mol H}^+ \text{ ha}^{-1} \text{ yr}^{-1}$  would be consumed, depending upon the source of N taken up by trees. If it originated from biological fixation of atmospheric N it would enter the system without any proton turnover. If it originated from  $\text{NO}_3^-$  input either through rain or through fertilizer, uptake would involve proton consumption. Ammonium fertilizer would conversely cause the production of protons. During the aggrading phase after clearfelling, the upper soil horizons tend to become more acidic (due to  $\text{NH}_4^+$  uptake and nitrification) and the lower horizons more basic due to  $\text{NO}_3^-$  uptake.

Harvesting interrupts the cycle of nutrient uptake and mineralization. The significance of biomass export on the  $\text{H}^+$  balance in the system depends on whether N is taken up as a  $\text{NH}_4^+$  or  $\text{NO}_3^-$  and the chemical composition of the biomass. As long as only the woody parts of the tree are exported from the ecosystem, the acidification effects will be negligible in mull soils with nitrification, or close to negligible in moder soils with mixed type and  $\text{NO}_3^-$  forms of N-nutrition but significant in mor type soils with an  $\text{NH}_4^+$  form of nutrition. If the N-rich parts are exported, the potential risk of acidification increases strongly. However, this will happen only if ion uptake included a surplus of cations, leading to the production of protons. Thus a high rate of production and utilization of woody parts under plantation systems need not be in conflict with the long-term stability of the ecosystem, provided the salts, including N, exported with the wood are replenished from time to time through fertilization and other means. However, it is important that the decomposer chain remains intact, including the link involving soil burrowing animals. The ecological consequences such as acidification caused by the export of biomass from a plantation are of prime importance, and the actual nutrient export may be only of secondary importance, though significantly

more effort has gone in studying the latter (e.g. Wells and Jorgensen, 1979). 1979).

#### B. Podzolization and other Effects of Plantation Trees on Soils

The numbers and species diversity of native soil fauna commonly tends to decrease in soils which have been planted to exotic tree species (Cornaby *et al.*, 1975). Such a decline, especially marked in the case of burrowing animals, probably reflects changes in the animal's environment and food, closely related to changes in the chemical status of the soil. The lack of mixing in such soils, which are usually passing through a phase of humus disintegration, means that each horizon of the soil should have a coupled element cycle (cation/anion) balance of its own order to check the acidification of its horizons. However, this is rarely possible and a partial decoupling of mineralization (proton consumption due to  $\text{NH}_4^+$  formation in the top organic layer) from ion uptake (proton production due to  $\text{NH}_4^+$  uptake) in the mineral soil, may lead to the acidification and podzolization of mineral soil.

Podzolization is usually associated with the diminished decomposability of the litter, which is probably due to a change in the chemical composition of the litter. A reduction in the decomposition of litter leads to an acidification phase in the soil, primarily because an appreciable part of the nitrogen uptake is from an  $\text{NH}_4$  source in the surface mineral soil. The proton production associated with this process remains unbalanced if N is accumulated in the litter layer. Moreover, low decomposability of litter may be coupled with tolerance to Al toxicity through the production of water-soluble organic substances by the vegetation (Swift *et al.*, 1979).

In even-aged plantations the spatial and temporal decoupling of the ion cycle and thus of podzolization may be quite marked, especially if the species involved is Al-tolerant and produces a litter of low decomposability. Indications of its occurring even during single rotation have been reported in spruce (Zezschwitz *et al.*, 1973; Grieve, 1978) and *Pinus radiata* plantations (McIntosh, 1980) and an increased cumulative effect is expected after multiple rotations of these species.

#### VI. CONCLUSIONS

A forest plantation is a dynamic system aimed at maximizing wood production, without impairing the long-term nutrient supply of the soil. It is

therefore important to identify the soil characteristics which determine the supply of nutrients to trees, and are sensitive to the manipulation necessary to establish a plantation (see Chapter 14). The mineralization and reuse of nutrients are of particular importance.

The balancing of the uptake and mineralization processes is a step towards bringing stability to the ecosystem, whereas their decoupling may lead to the excessive leaching of nutrients and to the initiation of podzolization processes in soils. The extent of decoupling, which is due to a net production of protons, depends upon the nature of the soil buffering systems present in the plantation soil. It is important to identify the type of buffering system involved and the process through which the decoupling of mineralization and nutrient uptake has occurred. The following silvicultural practices may assist in minimizing the decoupling:

- (1) The inter-rotational period should be kept short and a plant cover (possibly a legume) should be maintained during this phase. This will avoid the loss of nutrients due to leaching and erosion.
- (2) Leguminous crops may be used as intercropping or between rotations in plantations so as to increase the levels of N in the cycling fraction of the ecosystem, thereby improving the rate of litter decomposition and thus delaying the phase of humus (litter) accumulation. Addition of N fertilizers to N-deficient forest systems, especially at later ages, would have effects similar to those of growing a leguminous crop. The form of fertilizer N would however determine whether the protons are consumed or produced in the soil; an ammoniacal fertilizer causing acidity at the site of its uptake or nitrification.
- (3) With nutrient-deficient systems on acid soils, an application of lime and P will help to correct acidity and to create a more favourable C/P ratio. This will increase the rates of litter decomposition.
- (4) The provision of suitable substrate conditions for maintaining active populations of soil and litter fauna should be explored. This may include the retention of logging slash and litter (crushing/mulching slash), the use of leguminous crops, or the use of a mixture of two or more tree species in plantations. An application of lime and P may also be needed under certain soil conditions.
- (5) The likely benefits resulting from decomposition of slash and accumulated litter should be considered in relation to thinning operations.
- (6) Methods of harvesting and site preparation which do not lead to excessive losses of nutrients, either in biomass, in volatile forms or via transport by water should be selected.

Nutrients undergo transformation in their chemical nature as a result of ecosystem processes, and the significance of management practices to tree

nutrition will be reflected by changes in the chemical nature and rate of nutrient supplying processes in forest floors and soil layers. It seems important therefore, that studies in tree nutrition should aim to identify the most important forms of nutrients and rate processes, even though such studies are difficult and expensive to carry out. A practical solution of this problem may be to establish the relationships between easily obtainable soil parameters in the field and the nature and rate supply of various forms of important nutrients. So far preliminary attempts have been made in this direction; e.g. Shrivastva and Ulrich (1977) give examples of such relationships for *Picea abies* plantations in Hesse State of West Germany.

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